Modern and last glacial maximum snowline in Peru and Bolivia: implications for regional climatic change

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Modern and last glacial maximum (LGM) snowlines in the central Andes (5°-23° S) have been mapped using remote sensing techniques and geographical information system technology. The general configuration of the snowline during the LGM was similar to present with the snowline rising from east to west in response to decreasing precipitation. LGM snowline depression in the region deviates considerably from the 1000 m depression often assumed for low latitudes. A snowline depression model (Kuhn 1989) was used to investigate the temperature and precipitation changes necessary to cause the observed LGM snowline depression. Increased precipitation during the LGM is required to explain a portion of the observed 800-1200 m depression in the western Cordillera. Snowline depression of ≥ 1200 m along the eastern Andean slopes is the best proxy for the temperature depression experienced in the region during the LGM and can be explained by a cooling of approximately 5 to 7.5°C.

El límite de nieve perenne actual y la correspondiente a la de la última máxima glaciación (UMG) de los Andes Centrales (5°-23° S) han sido mapeadas utilizando técnicas de sensores remotos y sistema de información geográfica. La configuración general del límite de nieve perenne durante la UMG era similar a la actual, elevándose de este a oeste en respuesta a la disminución de las precipitaciones. La depresión del límite de nieve perenne durante la UMG difiere considerablemente de los 1000 m generalmente asumidos para bajas latitudes. Se utilizó el modelo de depresión del límite de nieve perenne (Kuhn 1989) para investigar los cambios de temperatura y precipitación necesarios para causar la depresión del límite de nieve perenne durante la UMG. Se requiere un aumento en la precipitación durante la UMG para explicar la depresión de 800-1200 m de la Cordillera Occidental. La depresión ≥ 1200 m del límite de nieve perenne a lo largo de las pendientes orientales andinas es la mejor aproximación para explicar la disminución de la temperatura ocurrida en esta región durante la UMG y puede ser explicada mediante un enfriamiento aproximado entre 5 y 7.5 °C.

Dans les Andes centrales (5°-23°S), le front actuel des neiges éternelles et celui datant du dernier maximum glaciaire (DMG ou LMG) ont étécartographiés par télédétection et par un système utilisant une technologiebasée sur des informations géographiques. La configuration générale dufront des neiges éternelles du DMG est semblable à la configurationactuelle. Ce front s'élève d'est en ouest suivant des précipitations décroissantes. La limite des neiges éternelles du DMG dans la régions'écarte considérablement des 1000m souvent rencontrés dans les zones debasse latitude. Un modèle décrivant l'abaissement du front des neigeséternelles (Kuhn, 1989) a été utilisé afin de déterminer les changements detempératures et de précipitations responsables de l'abaissement du frontdes neiges DMG. L'abaissement du front des neiges éternelles à

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800-1200m dans la cordillère occidentale durant le DMG s'explique en partie par une augmentation des précipitations. Sur les flancs de la cordillère orientale,l'abaissement du front neigeux supérieur à 1200m est ce qui révèle le mieuxle refroidissement subi par la région pendant le DMG. Il correspond à unebaisse d'environ 5 à 7,5°C.

Introduction

The geomorphology and stratigraphy of the central Andes record at least two major late Pleistocene glacial advances, one of which occurred before 20 ka BP and the other just prior to deglaciation at 14 ka BP (Seltzer, 1990, 1994a). These advances left a clear mark on the landscape in the form of cirques, moraines, and U-shaped cross-valley profiles that are well suited for mapping at regional scales using remote sensing techniques. This mapping allowed determination of the elevation difference between modern glaciers and those of the late Pleistocene glacial maxima on a regional scale to which we refer collectively as the LGM snowline depression.

Snowline depression can be theoretically linked to climatic perturbations (Kuhn, 1989; Ohmura *et al.*, 1992) and has been used by numerous authors to infer climatic change in the central Andes. However, most of these studies suffer from two major shortcomings: (1) the limited number of snowline observations that form the basis of the paleoclimatic interpretations and (2) extrapolation of results from a single site to a larger region. We addressed these shortcomings by compiling a detailed picture of the snowline elevation at present and during the LGM between 5° and 23° S. The compilation is based upon thousands of snowline measurements and was accomplished by combining previously published information (Nogami, 1976; Jordan, 1991; Fox, 1993; Fox and Bloom, 1994) with additional mapping of modern and LGM snowlines based on Landsat Thematic Mapper (TM) imagery in a geographical information system (GIS). Coupling of this mapping with knowledge of present climatic variables and a snowline depression model enabled us to relate the observed snowline depression to changes in temperature and precipitation.

<u>Methodology</u>

A comprehensive mapping of the present state of central Andean perennial snow and glacier cover (Figure 1) and the maximum limit of glacier extent during the late Pleistocene (Figure 2) served as the foundation of this study. In order to undertake such comprehensive mapping two simplifying assumptions were necessary. First, all the geomorphic features used to define the past state of glaciation are assumed to form contemporaneously during what we refer to as the last glacial maximum or LGM. The timing of the LGM is assumed to be coincident with time period represented by the CLIMAP (1984) reconstructed sea-surface temperatures. The validity of this assumption is discussed in Seltzer (1994b). Secondly, several different methodologies are employed to determine snowline elevation which necessitated the assumption that all of the methods yield equivalent snowline elevations. While not strictly true, the average elevational difference among the various methods employed here is only approximately 200 m. Some 14,000 points were used to define the modern snowline and 21,800 points were used to infer the LGM snowline. Regional snowline maps were constructed by averaging elevations within 15' latitude x 15' longitude cells to eliminate local variability. The averaged elevations were then contoured to produce the snowline maps. The procedures used in the snowline compilations are described below.

The modern snowline is defined to be the orographic snowline, which is the lower limit of perennial snowcover (Flint 1971 p. 64). In Peru, the modern snowline was taken as the lower limit of modern snowcover recorded on the Peruvian 1:100,000 map series published by the Instituto Geografico Militar, Lima, Peru (Fox, 1993; Fox and Bloom, 1994). Snowline elevations were recorded at distances of 1 to 2 km along the lower edge of snowcover. Construction of the regional snowline utilized the mean rather than the minimum in each 15' x 15' cell to minimize the variability caused by the effect of non-permanent snow. In southern Peru, Bolivia, and northern Chile, a similar approach was taken utilizing TM imagery rather than topographic maps. Computer classification was used for snowcover determination, allowing an objective assessment of snowcovered areas. We rejected elevations of probable non-permanent snowcover shown on the satellite imagery. In the eastern Cordilleras of Bolivia, the glacial inventory of Jordan (1991) was also used. Although the Equilibrium Line Altitudes (ELAs) used in Jordan's survey are a better estimation of the modern snowline altitude than the orographic snowline used here and by others, our orographic snowline is lower on average than the ELAs by only a small mean distance of 159 m in the Cordillera Real and Tres Cruces (Quimsa Cruz) regions for the 15' x 15' cells in which both measurements were made.

As with the modern snowline, two methods were employed to determine the LGM snowline in the central Andes. Over most of Peru, comprehensive TM coverage was not available and cirque floor altitudes were used as an approximation of the regional snowline (Fox and Bloom 1994). In Bolivia, Chile, and portions of Peru where TM coverage was available, the former extent of each glacier was mapped using the position of a terminal moraine, or in a few valleys, an abrupt change in valley form. To calculate a snowline altitude that is comparable to the cirque floors used in Peru the toe-to-headwall THAR method (Meierding 1982) was used. The toe (minimum) and headwall (maximum) elevation for each glacier were determined and the ELA was calculated as 0.45 of the vertical distance from the toe to the headwall (i.e. it has a toe-to-headwall altitude ratio or THAR value of 0.45). The selection of a THAR of 0.45 was based upon analysis of the THARs calculated for the modern glaciers mapped by Jordan (1991). Adjustment of the THAR ratio by ±0.10 caused an average elevational change of only ±50 m in our dataset.

From the regional snowline elevation of both the modern and LGM snowlines, LGM snowline depression was calculated by simply subtracting the two contoured surfaces. The resulting surface was then smoothed with a low-pass filter to produce the snowline depression map (Figure 3).

Snowline Observations

Both the modern and LGM snowlines increase in elevation from east to west across the Andes. The modern snowline rises from as low as 4300 to 4400 m on the eastern Andes in Bolivia to ≥ 5800 m in southwest Bolivia. The LGM snowline rises from between 3200 and 3600 m along the eastern slopes in Peru and Bolivia to 4900 m in southwest Bolivia. A similar gradient has been described previously (Satoh, 1979; Fox and Bloom, 1994; Seltzer, 1994b). Because the modern snowline rise closely parallels the east to west precipitation decrease, snowline elevation exhibits a strong response to decreasing precipitation. The east to west rise of both the modern and LGM snowlines demonstrates that the source of precipitation to the glaciers in the central Andes during the LGM was generally similar to that at present, the Amazon Basin.

The most important result of this study is that the assumption of a uniform 1000 m snowline depression the tropics during the LGM (Broecker and Denton 1989) is an oversimplification of the actual depression observed in the central Andes. For instance, areas in southeastern Peru experienced only 500 m of LGM snowline depression. Snowline depression was greatest along the humid eastern cordilleras where it consistently equaled or exceeded 1200 m. A small region of snowline depression in excess of 1200 m occurred on the western cordillera in northern Chile. While some areas of the central Andes, in particular the eastern slopes of the Andes, experienced a snowline lowering of 1000 m or more, this cannot be considered typical for most of the Altiplano.

Paleoclimatic Interpretations

Lowering of snowline elevations during the LGM were the result of a changing glacial mass balance in response to perturbations in temperature, precipitation, and radiation (Kuhn 1989, Ohmura 1992). The snowline depression in the central Andes during the LGM could be due solely to a perturbation in one of these factors, but more likely was caused by simultaneous perturbations in all three. We investigated lowering of snowline elevation in the context of the snowline depression model of Kuhn (1989), focusing on the effects of temperature and precipitation. At the altitude of the glacial equilibrium line, which is inferred to be the altitude of the snowline, the mass balance is zero and annual net accumulation equals annual net ablation. Changes in the equilibrium line altitude, or in this case snowline depression, can be calculated as changes in these quantities. In this model (equation 1), snowline depression (Δh) was calculated as a function of the length of the ablation season (τ), the latent heat of fusion

(L), and perturbations and vertical gradients, respectively, of temperature (δT_a, $\partial T_a/\partial z$), accumulation (δc, $\partial c/\partial z$) and net radiation (δQ_r, $\partial Q_r/\partial z$). We also note that this equation assumes that melting is the only form of ablation.

$$\Delta \mathbf{h} = \frac{\frac{\tau}{\mathbf{L}}\left(\delta \mathbf{Q}_r + \alpha(\delta \mathbf{T}_a)\right) - \delta \mathbf{c}}{\frac{\partial \mathbf{c}}{\partial \mathbf{z}} - \frac{\tau}{\mathbf{L}}\left(\frac{\partial \mathbf{Q}_R}{\partial \mathbf{z}} + \alpha\left(\frac{\partial \mathbf{T}_a}{\partial \mathbf{z}}\right)\right)} \tag{1}$$

Monthly mean climate data from the Goddard Earth Observing System Office General Circulation Model (GEOS-1) for 1986 are used in this study. GEOS-1 is a multi-year global atmospheric data set on a 2.5° x 2.0° grid and with 18 vertical pressure levels in the atmosphere (Schubert *et al.* 1993). Values for t, $\partial T_a/\partial z$, and $\partial Q_r/\partial z$ were derived from these data and $\partial c/\partial z$ and α (a bulk turbulent transfer coefficient for sensible heat) were taken from Kuhn (1979, 1989). The GEOS-1 assimilation model data show that considerable variations in τ, $\partial T_a/\partial z$, and $\partial Q_r/\partial z$ occur throughout the central Andes. The values used in the following discussion (listed in Table 1) represent the calculated average values of these variables for the entire region.

Table 1: Values used for parameterizing Kuhn's (1989) snowline model for the central Andes.

variable	value	units
τ	180	day
$\partial T/\partial z$	-0.006	°C m^{-1}
$\partial Q_r/\partial z$	0.001	MJ day^{-1} m^{-1} m^{-2}
α	1.7	MJ day^{-1}
$\partial c/\partial z$	1	mm m^{-1}
δT	variable	°C
δQ_r	0	MJ day^{-1} m^{-2}
δc	0	mm
L	330000	J kg^{-1}

This simple snowline depression model, coupled with the observed LGM snowline depression (Figure 3), is used to address two paleoclimatic questions. The first concerns the effect of precipitation on snowline depression in the central Andes during the LGM, which has been the source of disagreement. Hastenrath (1967, 1971) and later Wright (1983) argued that increased precipitation is the primary cause of LGM snowline depression in the arid western Cordillera. However, Fox and Bloom (1994) argued the opposite view; snowline lowering was caused solely by a temperature decrease. In fact, they argued that precipitation decreased over much of the Altiplano during the LGM thereby causing less snowline depression than would otherwise have occurred. The second question we address is reconciling the consistent □ 1200 m snowline depression observed in the eastern Cordillera with a relatively minor (2 - 3°C) decrease inferred for low-latitude sea-surface temperature during the LGM (CLIMAP 1981).

In the tropics and subtropics, season variations in temperature are small and the elevation of the glacier equilibrium line is primarily controlled by the elevation of the 0° isotherm where precipitation is not a limiting factor. As snowline rises above the elevation of the zero degree isotherm in response to decreasing precipitation, the duration of the ablation season (τ) decreases. From equation (1) it can be seen that a decreasing t will diminish the effects that temperature and net radiation changes will have on snowline lowering. In the limiting case where the duration of the ablation season is zero, temperature and net radiation changes have no effect on snowline lowering and snowline will only respond to changes in accumulation. In the western Cordilleras where snowlines are well above the annual 0° isotherm, precipitation increases will have a large effect of snowline lowering. This observation is not new and has been pointed out previously (e.g. Hastenrath 1967, 1971). Equally important, and often overlooked, is the observation that a temperature decrease will have the greatest effect on snowline lowering in the eastern Cordilleras of Peru and Bolivia where snowlines are the lowest and melt duration is the longest.

This simple analysis has two important implications. First, to invert a snowline depression to a temperature depression, it is best to use the snowline depression observed along the eastern Cordilleras where the ablation season (t) is the longest. The longer the ablation period, the smaller is the effect that uncertainties in knowledge of t will have on the estimated temperature change. Second, if the ablation season is indeed very short at high elevations, as the GEOS-1 data indicates, then temperature alone cannot be responsible for the 800-1000 m LGM snowline depression found along the western Cordillera in southern Peru and northern Chile.

To investigate how a given cooling will impact snowline lowering as a function of the length of the ablation season, the effect of a 7.78° C cooling on snowline depression was calculated for varying ablation duration (Table 2). It is clearly the case that as the duration of ablation season decreases, the magnitude of the snowline depression also decreases dramatically. We add that this simple approach neglects sublimation, an important ablation process in the arid western Cordillera. If sublimation becomes an important ablation processes then the energy available for melting is dramatically reduced and the sensitivity of a snowline depression to a temperature cooling is further decreased. This simple analysis suggests that LGM snowline depressions in areas of the central Andes where snowlines are well above the

0° isotherm must have been caused by increased precipitation during the LGM. It is unlikely that snowline would have experienced a ~1000 m depression under the dramatically drier (80% ±25%) conditions for the LGM hypothesized by Fox and Bloom (1994). Because of present uncertainties in the actual duration of the ablation season (τ), the vertical gradient of accumulation ($\partial c/\partial z$), and the importance of sublimation, quantitative estimates of the precipitation changes necessary to produce the observed snowline depression are not attempted.

Table 2: Effect of varying duration of ablation season (Δ) on snowline depression (Δh) for a cooling (ΔTa) of -7.78 °C. All other variables are as listed in Table 1.

Δ (days)	0	15	30	45	60	90
Δh (m)	0	-424	-655	-800	-900	-1028

As discussed above, to assess potential temperature depressions during the LGM accurately, the ≥ 1200 m snowline depression observed along the eastern Cordilleras should be used. Solving equation (1) for δT_a using the parameterizations from Table 1 reveals that a 7.78 °C cooling is sufficient to cause the observed snowline depression. This is in agreement with the 6.6-8.4 °C cooling inferred by numerous authors (including ourselves) by simply multiplying the 1200-1400 m snowline depression with the observed atmospheric lapse rates of 5.5 to 6.0 °C km^{-1}.

However, the modeled temperature change is strongly dependent on the assumed vertical gradient of accumulation ($\partial c/\partial z$). In the analysis so far we have assumed $\partial c/\partial z$ to be 1 mm m^{-1}, an average value for conditions encountered in the Alps (Kuhn 1989).

Several lines of evidence suggest, however, that this may not be representative of conditions on the eastern slopes of the Andes. In the central Andes regions of maximum precipitation occur on the eastern slopes at elevations much lower than present snowline. Ribstein *et al.* (1995) also found a slightly negative precipitation gradient at elevations 3500-4700 m near the Zongo Glacier in the Cordillera Real, Bolivia. If during LGM times, the precipitation that presently falls as rain had been converted to snow, our assumed accumulation gradient in these areas would be too high, and in fact could be of the wrong sign. Satoh (1979) also suggested that during the LGM maximum precipitation would have been on the eastern midslopes rather than at the highest elevations. The effect of lowering the accumulation gradient was tested by decreasing $\partial c/\partial z$ from 1.0 to -1.0 mm m^{-1} in at various increments and recalculating the temperature change required to produce the observed depression (Table 3). If the vertical gradient of accumulation ($\partial c/\partial z$) was near or slightly negative along the eastern Cordilleras during the LGM, the modeled temperature depression required to explain the observed 1200 meter snowline depression is between 5 to 7.5 °C.

Table 3: Effect of changing accumulation gradient (∂c/∂z) on modeled temperature depression(δT_a) for a duration of melting (τ) of 90 and 180 days. All other variables are as listed in Table 1.

∂c/∂z (mm m^{-1})	**temperature depression (°C) τ=90 days**	**temperature depression (°C) τ=180 days**
1.0	9.0	7.8
0.5	7.8	7.14
0.25	7.1	6.87
0.1	6.6	6.62
0	6.75	6.49
-0.1	6.23	6.36
-0.25	5.9	6.17
-0.5	5.2	5.85
-1.0	3.9	5.2

Conclusions

Remote sensing and geographical information systems (GIS) technologies enabled construction of detailed modern and LGM snowline maps and calculation of the LGM snowline depression in the central Andes. The general configuration of the snowline during the LGM was similar to that of the present. Both snowlines show a strong rise from east to west in response to decreasing precipitation across the Andes. LGM snowline depression over much of the Altiplano was much less than the 1000 m often assumed for low latitudes. The short ablation season duration in the western Cordillera, especially in southwestern Bolivia, requires that snowline lowering was caused in part by increased precipitation during the LGM. The magnitude of the increase is not presently known. Snowline depression on the humid eastern margin of the Andes is the best proxy for the temperature depression experienced in the region. A LGM temperature depression of approximately 5 to 7.5 °C is adequate to explain the observed eastern snowline depression of ≥1200 m.

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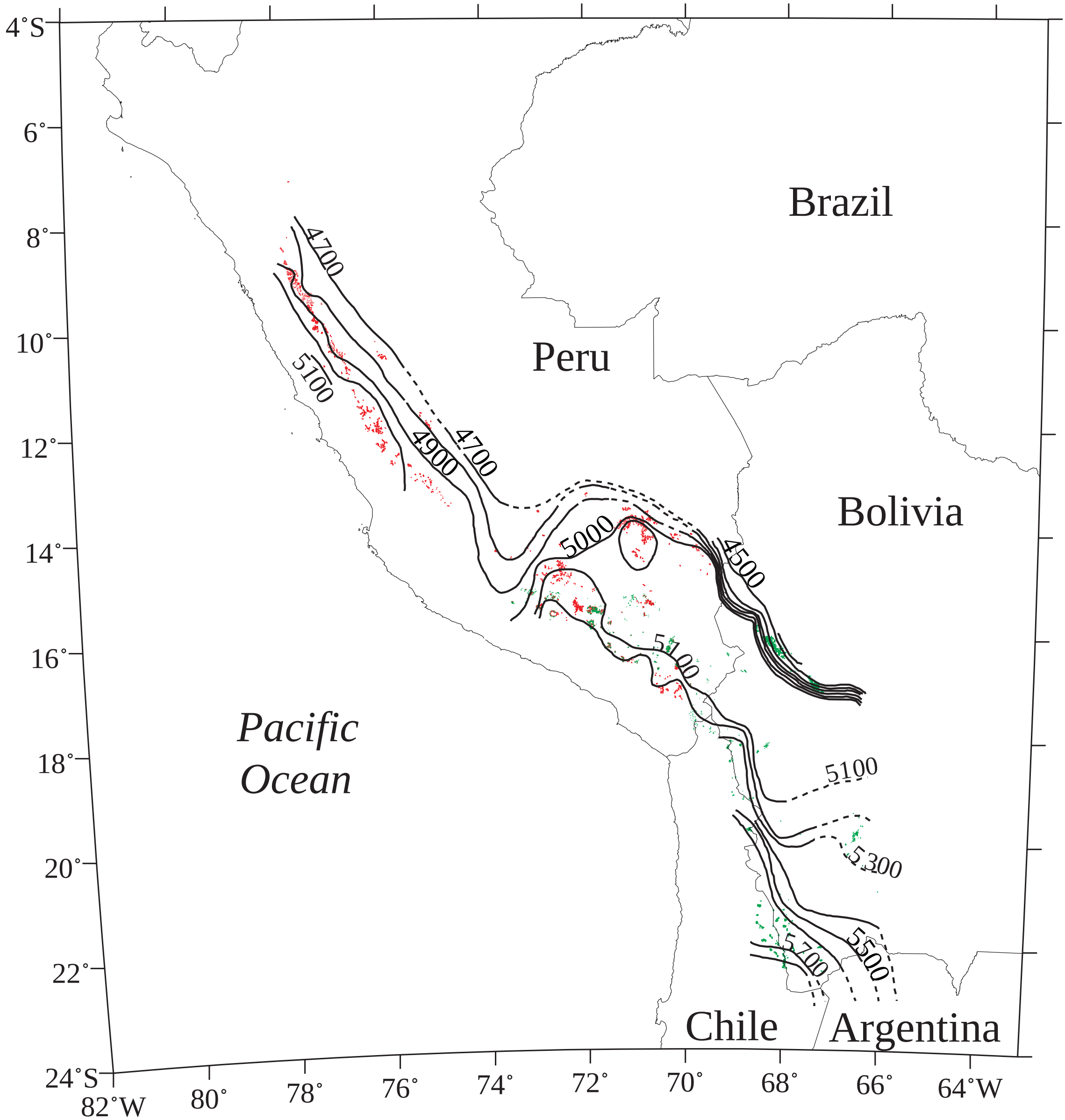

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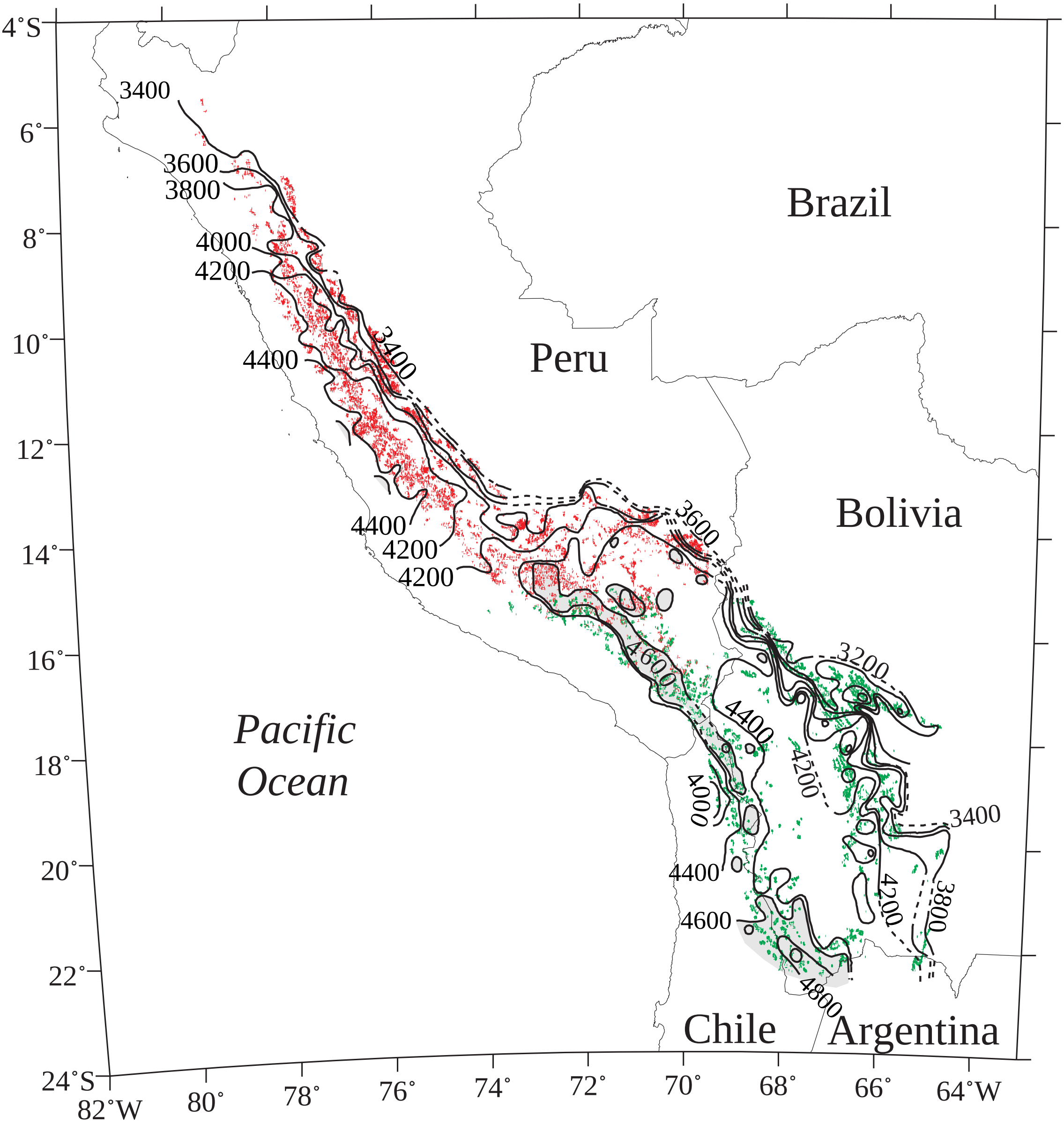

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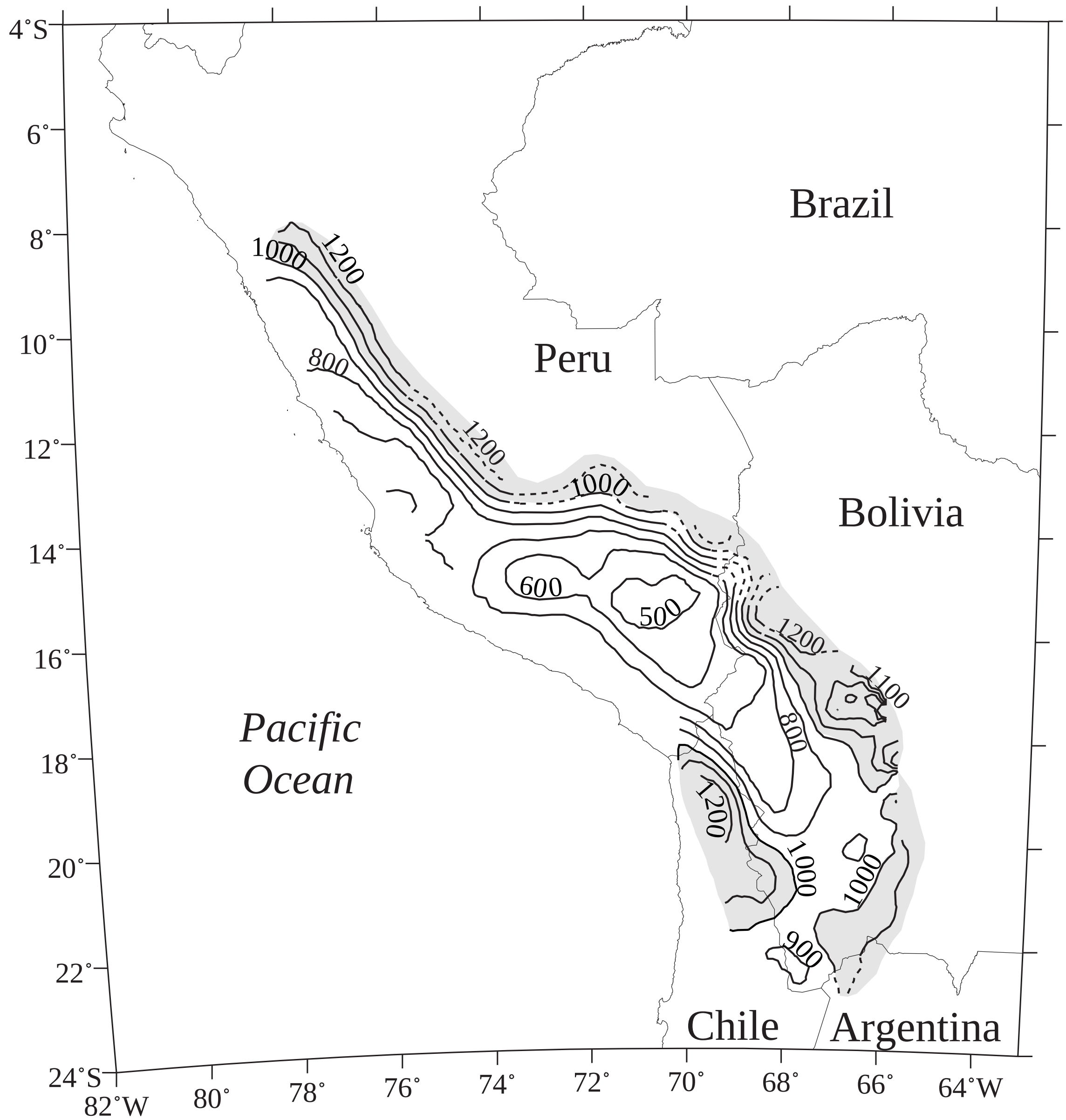

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